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1 **Timescales of successful and failed subduction: insights from numerical modelling**

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10 Abstract

11 The relatively short duration of the early stages of subduction results in a poor geological record,
12 limiting our understanding of this critical stage. Here, we utilize a 2D numerical model of incipient
13 subduction, that is the stage after a plate margin has formed with a slab tip that extends to a shallow depth
14 and address the conditions under which subduction continues or fails. We assess energy budgets during
15 the evolution from incipient subduction to either a failed or successful state, showing how the growth of
16 potential energy, and slab pull, is resisted by the viscous dissipation within the lithosphere and the mantle.
17 The role of rheology is also investigated, as deformation mechanisms operating in the crust and mantle
18 facilitate subduction. In all models, the onset of subduction is characterized by high lithospheric viscous
19 dissipation and low convergence velocities, whilst successful subduction sees the mantle become the main
20 area of viscous dissipation. In contrast, failed subduction is defined by the lithospheric viscous dissipation
21 exceeding the lithospheric potential energy release rate and velocities tend towards zero. We show that
22 development of a subduction zone depends on the convergence rate, required to overcome thermal
23 diffusion and to localise deformation along the margin. The results propose a minimum convergence rate
24 of $\sim 0.5 \text{ cm yr}^{-1}$ is required to reach a successful state, with 100 km of convergence over 20 Myr,
25 emphasizing the critical role of the incipient stage.

26 **Key words:** Subduction zone processes, Rheology: mantle, Dynamics of lithosphere and mantle

1. Introduction

Subduction plays a key role in Earth's evolution as it provides the main driving force for plate motions, on-going deformation and the cycling of material in and out of the mantle [Forsyth and Uyeda, 1975]. The critical stages in the development of subduction zones are the initiation, where the plate margin is formed, and the incipient phase, where the slab tip extends to a shallow depth. This is followed by the lengthening of the slab at depth, until subduction eventually becomes self-sustaining. While self-sustained subduction has been subject of many studies [e.g. Conrad and Hager 1999; Funiciello et al. 2003; Billen and Hirth 2007; Capitanio et al. 2007; Capitanio et al. 2009; Schellart 2009; Garel et al. 2014], the incipient stage is poorly understood. This is partly because of the short duration of the incipient stage, and the record of this stage may be overprinted once the margin matures [Gurnis et al., 2004; Stern and Gerya, 2018]. This has hence made subduction initiation and the incipient stage, a difficult process to observe directly and therefore understand.

Subduction is the result of a gravitational instability, requiring two main conditions to occur: (1) deformation of the lithosphere to form a plate margin and (2) lengthening of the slab, to grow the gravitational instability and attain a self-sustaining state. Previous work addressing plate margin formation has focussed on the yield strength of lithosphere, which has to be overcome. The timescales for a new plate margin to form range from 2 to 5 Myr during forced convergence at pre-existing weaknesses [Hall et al., 2003; Gurnis et al., 2004] and may be aided by rapid plastic failure due to the addition of water and elasticity [Regenauer-Lieb et al., 2001]. Although very limited field data exists, observations from the Semail Ophiolite suggest subduction reached a self-sustaining state after ~ 8 Myr of forced convergence in the region [Guilmette et al., 2018].

Previous modelling studies, both numerical and analogue, have addressed subduction assuming an initial configuration with a slab extending to a depth of 100 to 250 km [e.g. Capitanio et al., 2010;

MacDougall et al., 2017; Agrusta et al., 2017] whilst force balance from McKenzie [1977] suggests a critical depth of 180 km for self-sustaining subduction at a minimum convergence rate of 1.3 cm yr^{-1} . Hall et al. [2003] and Gurnis et al. [2004] suggest 100 – 150 km of convergence is required along a fracture zone at a rate of $\geq 1 \text{ cm yr}^{-1}$ to achieve self-sustaining subduction. Even with imposed boundary velocities, the growth of the instability is controlled by the deformation rate of the lithosphere and mantle. The slab begins to sink at low velocities initially, before exponentially increasing as resisting forces are overcome and slab pull becomes the dominant driving force, leading to self-sustaining subduction [Becker et al., 1999; Gurnis et al., 2004]. Additionally, the rheology of the surrounding mantle plays a role in the development of subduction zones, reducing the viscous drag at the slab edge [Gurnis et al., 2004; Billen and Hirth, 2005; Billen and Hirth, 2007; Jadamec, 2015]. Gurnis et al. [2004] showed that the strength of the mantle influences the evolution of the margin once the mantle viscosity exceeds 10^{20} Pa s , and only after the lithospheric strength has been overcome initially.

These studies suggest conditions for successful subduction, thereby defining the conditions under which subduction fails. To develop into a successful subduction margin, the negative buoyancy of the slab penetrating into the mantle must remain high enough to drive subduction. In a thermo-mechanic system, the nascent buoyancy force, during the incipient stage, may remain comparable or smaller than the resisting forces, resulting in a decrease of the deformation rate of the lithosphere and subduction velocity. Under these conditions, thermal diffusion is favoured and, as a consequence, the slab's temperature anomaly and its thermal buoyancy may vanish. This negative feedback may rapidly lead to subduction failure.

It is known that the evolution of subduction follows the balance of forces, with driving forces including slab pull (F_{SP}) and ridge push (F_{RP}), and induced mantle flow (F_{S}) [Baes and Sobolev, 2017; Baes et al., 2018]. Resisting forces include the slab strength, resulting in a resistance to bending prior to

73 the trench (F_B), plate coupling at the plate boundary and its resistance to shearing (F_{PC}), the surrounding
74 mantle's viscous resistance through drag (F_D) and by the amount of over-riding plate deformation (F_{OPD})
75 [McKenzie, 1977; Toth and Gurnis, 1998; Capitanio et al., 2007]. During incipient and self-sustaining
76 subduction, the evolution of the slab is determined by the energy dissipation and its partitioning between
77 the lithosphere and mantle. Previous studies, using both analogue and numerical modelling techniques,
78 have illustrated how the energy in a subducting system is dissipated during self-sustaining subduction.
79 Viscous dissipation in the lithosphere, as a percentage of the total energy, during self-sustaining
80 subduction with no over-riding plate in both analogue and numerical models has been estimated to range
81 from less than 20% [Stegman et al., 2006; Capitanio et al., 2007; Capitanio et al., 2009; Irvine and
82 Schellart, 2012; Schellart, 2009] to around 50% [Buffett, 2006; Di Giuseppe et al., 2008] and up to as high
83 as 95% [Conrad and Hager, 1999; Funiciello et al., 2003; Bellahsen et al., 2005]. A global reconstruction
84 of energy dissipation at subduction zones by Stadler et al. [2010], estimated lithospheric dissipation at less
85 than 25%, with the lithospheric component almost equally distributed between the weak zone and hinge
86 region. Elasticity has been shown to maintain higher deformation rates during bending and retain slab
87 strength during unbending but has minor effects on the overall morphology [Capitanio and Morra, 2012;
88 Farrington et al., 2014].

89 This study investigates the evolution energy dissipation during the incipient subduction stage using a
90 numerical model to determine the conditions for the growth of an incipient slab. The incipient stage is
91 addressed here as the initial deformation phase that is focused within the lithosphere. In such a system,
92 the deformation of the lithosphere must be overcome before the slab can lengthen and induce mantle flow,
93 which leads to self-sustaining, successful subduction. We investigate how the deformation is partitioned
94 between the lithosphere and mantle during the incipient stage through viscous dissipation and define the
95 timescale of the incipient stage. Additionally, key factors of the subduction system are addressed, such as

96 rheology and plate ages, to assess their influence on the evolution from incipient subduction, to either a
 97 successful or failed state.

98 2. Model Overview

99 Thermomechanical models of subduction are performed using the code Fluidity, which uses a
 100 finite-element, control-volume method and an adaptive, unstructured discretization [e.g., Davies et
 101 al., 2007; 2008]. The model dimensions are 10,000 km by 2,900 km, with node spacing varied from 400
 102 m to 200 km. The large dimensions reduce the influence of the boundary conditions [Enns et
 103 al., 2005; Chertova et al., 2012]. The higher resolution, which is driven by the adaptivity algorithm of
 104 Fluidity [Davies et al., 2011], is assigned to regions where variables experience a higher rate of change.

105 2.1 Governing equations and constitutive laws

106 We solve the conservation of mass, momentum and energy through the Boussinesq approximation
 107 of an incompressible Stokes fluid:

$$\partial_i u_i = 0, \quad (1)$$

$$\partial_i \sigma_{ij} = -\Delta \rho g_j, \quad (2)$$

$$\frac{\partial T}{\partial t} + u_i \partial_i T = \kappa \partial_i^2 T, \quad (3)$$

108 where u is the velocity vector and g is the acceleration due to gravity, σ_{ij} is the stress tensor due to
 109 strain rates. T is the temperature, κ the thermal diffusivity and $\Delta \rho$ is the density difference. The stress
 110 tensor, σ_{ij} , can be split into the deviatoric and lithostatic components:

$$\sigma_{ij} = \tau_{ij} - p\delta_{ij}, \quad (4)$$

111 where τ_{ij} is the deviatoric stress tensor, p is the dynamic pressure, and δ_{ij} is the Kronecker delta function.
 112 The deviatoric stress tensor and the strain rate tensor $\dot{\epsilon}_{ij}$ are related by equation (5), where η is the viscosity
 113 and subscript i and j represent the vector direction.

$$\tau_{ij} = 2\eta\dot{\epsilon}_{ij} = \left(\frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right) \quad (5)$$

114 2.2 Rheology

115 The model incorporates a viscoplastic rheology, where the plastic rheology is implemented through
 116 yielding at low pressure and temperature through a brittle-failure type yield-stress law (eq. 6 and 7) while
 117 a viscous rheology is implemented through a stress - strain rate constitutive law for each creep mechanism
 118 (eq. 8). All values for the rheology are outlined in Table 1. The plastic rheology is determined by:

$$\eta_y = \frac{\tau_y}{2\dot{\epsilon}_{II}} \quad (6)$$

119 where η_y the yielding viscosity and $\dot{\epsilon}_{II}$ the second invariant of the strain-rate tensor. τ_y , the yield strength,
 120 is calculated by:

$$\tau_y = \min(\tau_s + f_c P, \tau_{y,max}) \quad (7)$$

121 where τ_s is the surface yield strength, f_c is the friction coefficient, P is the lithostatic pressure and $\tau_{y,max}$
 122 is the maximum yield strength. The values prescribed for the friction coefficient along the weak zone and
 123 within the model are consistent with previous values [e.g. Cattin et al., 1997; Iaffaldano, 2012; Arcay,
 124 2012; Cramer et al., 2012; Cramer and Tackley, 2015]. Diffusion, dislocation and Peierls creep are
 125 determined by a constitutive law between stress and strain rate for each mechanism:

$$\eta_{diff/dis/P} = A^{-\frac{1}{n}} \exp \left(\frac{E + PV}{nRT_r} \right) \dot{\epsilon}_{II}^{\frac{1-n}{n}} \quad (8)$$

126 where A is the prefactor, n is the stress component, E the activation energy and V the activation volume,
 127 P the lithostatic pressure ($=\rho_s g z$), R the gas constant. T_r is the temperature obtained by adding to the
 128 Boussinesq approximation an adiabatic gradient of 0.5 K/km in the upper mantle and 0.3 K/km in the
 129 lower mantle [Fowler, 2005].

130 The effective viscosity (η) is then evaluated for the entire domain through:

$$\eta = \left(\frac{1}{\eta_{diff}} + \frac{1}{\eta_{dis}} + \frac{1}{\eta_P} + \frac{1}{\eta_y} \right)^{-1} \quad (9)$$

131 Viscosity is capped in the model with a lower limit at 10^{18} Pa s, and upper limit of 10^{25} Pa s. Diffusion
 132 creep dominates lower mantle deformation, with a viscosity contrast of 30 between the upper and lower
 133 mantle, consistent with seismic observations [Karato et al., 1995].

134 A 5 km thick weak zone is situated along the top of the subducting slab that also follows the same
 135 rheological laws as the upper mantle, however a lower friction coefficient is prescribed, along with a
 136 viscosity cap of 10^{20} Pa s, ensuring decoupling between the two plates. The interface between the two
 137 plates is free to move in response to the dynamic evolution of the model. This weak zone simulates weak
 138 hydrated crust on the top of the subducting slab, a critical condition for subduction to occur [Cramer et
 139 al., 2012] and the rheology of the weak zone allows strain rate weakening to occur within the weak zone
 140 so strain can naturally localise. This weak zone is phased out below a depth of 200 km [Garel et al., 2014].

141 2.3 Model setup

142 The model setup (Fig. 1A) is derived from the work of Garel et al. [2014], with the initial bending
 143 angle (β) and radius of curvature modified to investigate the boundary between successful and failed
 144 subduction. No heat flux occurs on the side walls, whilst isothermal conditions are set on the top (T_s) and

bottom (T_B) walls ($T_s = 273$ K, $T_B = 1573$ K). The model includes a free-slip boundary condition on the bottom and side walls, whilst a free surface is employed on the top surface to allow plate flexure in response to the underlying dynamics [Kramer et al., 2012].

The controlling parameter on the lithospheric strength is the temperature profiles of the subducting (SP) and over-riding plates (OP), dictated by the age-dependent thicknesses of the lithosphere. The thickness is determined via the half-space cooling model, where the age increases linearly from 0 Myr at the model boundary to the prescribed plate ages at the initial trench position at the centre of the domain ($x_{\text{trench}}^0 = 5,000$ km). The initial temperature profile is thus determined by:

$$T(x, z, t = 0) = T_s + (T_m - T_s) \operatorname{erf} \left(\frac{z}{2\sqrt{k \operatorname{Age}^0(x)}} \right) \quad (10)$$

where z is the depth, x the horizontal coordinate, t is the time, k is thermal diffusivity, T_s is the temperature at the surface and T_m the mantle temperature, with $\operatorname{Age}^0(x)$ is the plates age at the given x coordinate.

We investigated cases with different initial thermal structures. These are defined by SP and OP ages at the trench. For each plate age combination, we also investigated varying the initial geometry of the subducting plate, by either changing the bending radius, or the initial bending angle β . The initial parameters for each case are listed in Table 2, where the case name includes the SP and OP age in Myr and if the case reaches a successful case. In addition, we have also tested the role of rheology by varying the deformation mechanisms operating within the lithosphere and mantle and also by investigating the strength of the weak zone.

2.4 Diagnostics

The initial depth below the surface of the subducting slab is defined as the deepest point of the subducting slab tip, as defined by the 1300 K isotherm (Fig. 1A). The 1300 K isotherm is considered the base of the lithosphere, where $T > 1300$ K is regarded as mantle. The horizontal plate velocity of the

166 subducting (u_{SP}) and over-riding plate (u_{OP}) are measured at the surface where $x = 2000$ km and $x = 8000$
 167 km, respectively. u_{SP} and u_{OP} are used to determine the convergence velocity ($u_x = u_{SP} + u_{OP}$), where
 168 trenchward motion is positive [Garel et al., 2014]. The slab tip velocity (u_T) is calculated from the deepest
 169 point available of the 1300 K temperature contour which denotes the subducting slab.

170 The energy dissipation in the system is measured by the viscous dissipation (VD) through internal
 171 deformation [Ranalli and Murphy, 1987; Conrad and Hager, 1999; Capitanio et al., 2007]. The viscous
 172 dissipation is calculated across the volume (v) of the lithosphere (OP & SP combined, ϕ_l^{VD}), where $T <$
 173 1300 K, as well as the mantle (ϕ_m^{VD}), where $T \geq 1300$ K. The sum of the two is the total viscous dissipation
 174 in the system $\phi_l^{VD} + \phi_m^{VD} = \phi_{TOT}^{VD}$:

$$\phi_{TOT}^{VD} = 4 \int_V \eta \dot{\epsilon}_{II} \dot{\epsilon}_{II} dV \quad (11)$$

175 The system is primarily driven by the potential energy (PE) of the subducting slab, along with a
 176 minor component of ridge push. Additional potential energy in the system arises from the temperature
 177 contrast between the asthenosphere and adiabatic mantle. These are quantified through the lithospheric
 178 ϕ_l^{PE} and mantle ϕ_m^{PE} , PE terms respectively, the sum of which is the total release rate of potential energy
 179 in the system $\phi_l^{PE} + \phi_m^{PE} = \phi_{TOT}^{PE}$, which is determined by:

$$\phi_{TOT}^{PE} = \int_V \rho_s g \alpha [T_m - T] u_y dV \quad (12)$$

180 Where ρ_s is the density at the surface, α is the thermal expansivity, T is temperature and T_m is the mantle
 181 temperature, with u_y representing the vertical velocity [Conrad and Hager, 1999].

182 We find that $\phi_{TOT}^{PE} \approx \phi_{TOT}^{VD}$, with minimal energy within the system lost due to the free surface
 183 implementation. The viscous dissipation and potential energy release rate are presented as a percentage of
 184 the total ($\phi^{VD} = \frac{\phi_l^{VD}}{\phi_{TOT}^{VD}}$, $\phi^{PE} = \frac{\phi_l^{PE}}{\phi_{TOT}^{PE}}$) where the absolute values of each vary as a function of time. It is

185 important to note that a decrease in the lithospheric potential energy (ϕ_l^{PE}) over the total potential energy
186 release rate in the system (ϕ_{TOT}^{PE}) is the consequence of thermal diffusion. This allows us to identify key
187 information regarding the model, with viscous dissipation ϕ^{VD} used to identify the main area of resistance
188 within the system, whilst the potential energy release rate ϕ^{PE} is used to determine how slab develops,
189 whether it grows as the slab lengthens, or decreases as the slab thermally diffuses away.

190 Successful subduction is defined by the slab extending to mid upper-mantle depths and an increase in
191 convergence and sinking velocities, which indicates vertical sinking of the slab as it over-comes
192 lithospheric resistance. Failed subduction is identified through the vanishing of the slabs thermal signature
193 and convergence velocities in the model declining toward zero, as the instability is unable to grow.
194 Therefore, the combination of partitioning of energy within the system and convergence velocities best
195 characterise incipient subduction. The incipient stage is characterised by high lithospheric viscous
196 dissipation and low convergence velocities, which transitions to a decrease in lithospheric viscous
197 dissipation and an increase in convergence velocity in the successful stage. In contrast, failed subduction
198 is characterised by the proportion of lithospheric viscous dissipation exceeding the lithospheric potential
199 energy release rate and convergence velocities tending towards zero.

200

3. Results

201

3.1 Energy and dissipation partitioning evolution

202 We assess the evolution of energy and its dissipation within the system over time to determine the
203 transition between successful and failed subduction. In Table 2 we summarise the models' evolution
204 indicating the conditions under which subduction is successful and any modifications to deformation
205 mechanisms. The geometry and rheology of the slab are key components in the bouyancy-driven models.

206 To find the initial critical depth that results in successful subduction, we increased the depth of the slab
 207 tip (either through the bending radius or bending angle), to increase the slab pull force of the initial setup
 208 to facilitate subduction.

209 In case 20-100Y (Fig. 2A) the slab initially penetrates to a depth of 166 km. The lithospheric
 210 viscous dissipation ϕ^{VD} peaks at around 10 Myr at $\sim 75\%$, which coincides with the lowest convergence
 211 (u_x) and vertical (u_y) velocities (Fig. 3A, 3C). ϕ^{VD} continues to decline to $\sim 50\%$ after 42.5 Myr. The
 212 potential energy release rate from the lithosphere ϕ^{PE} reaches $\sim 75\%$ at 20 Myr where it then remains
 213 constant for the rest of the evolution. As the model evolves the absolute values of both viscous dissipation
 214 (ϕ_{TOT}^{VD}) and potential energy release rate (ϕ_{TOT}^{PE}) increase, whilst ϕ^{VD} decreases due to a decrease in the
 215 proportion of lithospheric viscous dissipation (ϕ_l^{VD}) and an increase in proportion of mantle viscous
 216 dissipation (ϕ_m^{VD}) occurs as slabs sinking velocity increases after the incipient stage and reaches a
 217 successful state.

218 The slab in case 20-100N (Fig. 2B) initially extends to 141 km below the surface, leading to failed
 219 subduction, with ϕ^{VD} exceeding ϕ^{PE} around 10 Myr. ϕ^{PE} also decreases from $> 80\%$ to $\sim 50\%$ over 50
 220 Myr, where the mantle component of potential energy release rate increases. In this case, the absolute
 221 values of both viscous dissipation and potential energy release rate decrease as the subduction zone fails,
 222 evident by velocities tending toward 0 cm yr⁻¹ (Fig 3B, D).

223 Case 100-100Y represents a slab that initially extends to a depth of 162 km, with an older
 224 subducting plate (100 Myr) compared with case 20-100Y (20 Myr SP). The peak in ϕ^{VD} , at $\sim 80\%$, occurs
 225 around 15 – 20 Myr, higher than any other successful case (Fig. 2C). However, the decline of ϕ^{VD} is faster
 226 than in case 20-100Y, resulting in a similar transition time from lithospheric to mantle dominated
 227 dissipation at ~ 42.5 Myr (Fig. 2A). ϕ^{PE} in case 100-100Y remains high at $> 80\%$, as the slab lengthens

(Fig. 2C). During case 100-100N, which results in failed subduction, (Fig. 2D), ϕ^{VD} exceeds ϕ^{PE} at around 6 Myr, shorter than any other failed case.

The slab in case 100-20Y (Fig. 2E) initially extends to a depth of 162 km and is unique as it is the only case where lithospheric dissipation remains high, with ϕ^{VD} at $\sim 70\%$ throughout the evolution of the model, remaining in an incipient state for > 80 Myr. ϕ^{PE} remains high at $\sim 80\%$, allowing subduction to continue, albeit at very low rates. In 100-20N, where subduction fails (Fig. 2F), the slab initially extends to a depth of 145 km, with ϕ^{VD} exceeding ϕ^{PE} after ~ 18 Myr.

Case 20-20Y (Fig. 2G) is also unique, as slab break-off is observed. ϕ^{VD} reaches a peak at around 10 Myr and then reaches the transition from lithosphere- to mantle-dominated viscous energy dissipation at around 17.5 Myr. ϕ^{VD} starts high at 60%, before decreasing to 10% at 22.5 Myr as the broken slab reaches a high vertical velocity (u_y , eq. 12, Fig. 3A) increasing strain rates in the mantle ($\dot{\epsilon}_{II}$, eq. 11). The value of ϕ^{VD} then increases to 40% at 35 Myr as subduction continues, with ϕ^{VD} then decreasing to $\sim 25\%$ after 60 Myr as the slab reaches the 660 km transition zone. ϕ^{PE} follows a similar trend, where it decreases from $> 80\%$ to $\sim 30\%$ at 22.5 Myr during slab break-off, before increasing to 80% at 45 Myr as subduction continues. In case 20-20N (Fig. 2H), the slab initially extends to a depth of 141 km below the surface. In 20-20N, the instability is unable to grow, ultimately leading to the proportion of ϕ^{VD} exceeding ϕ^{PE} after ~ 10 Myr.

The slab in case 65-40Y (Fig. 2I) initially extends to a depth of 185 km. In case 65-40Y, ϕ^{VD} reaches a peak between 10 - 15 Myr, before decreasing to $\sim 40\%$ after 27.5 Myr. The ϕ^{PE} in case 65-40Y remains relatively constant, at around 80% for the duration of the model. Velocities reach a minimum around 10-15 Myr, before increasing to 8 cm yr^{-1} at 27.5 Myr. Case 65-40N (Fig. 2J) initially extends to a depth of 154 km. ϕ^{VD} reaches a peak at 20 Myr, where it intersects with ϕ^{PE} . ϕ^{VD} exceeds ϕ^{PE} at

250 around 22.5 Myr, resulting in failed subduction. As the slab is unable to lengthen, velocities tend toward
 251 0 cm yr⁻¹ (Fig. 3B, 3D).

252 In case 40-65Y (Fig. 2K), the slab tip initially extends to 150 km, with a peak in ϕ^{VD} observed at
 253 10 – 15 Myr at ~ 70%. This peak corresponds to the lowest convergence velocity observed of 0.37 cm yr⁻¹.
 254 ¹. After the peak, ϕ^{VD} decreases to ~ 50 % and convergence velocity increases to 1.9 cm yr⁻¹ at 37.5 Myr.
 255 ϕ^{PE} remains high, at ~ 80% for the entire evolution. In case 40-65N, ϕ^{VD} exceeds ϕ^{PE} at ~ 12 Myr (Fig.
 256 2L), whilst velocities decrease from ~ 0.5 cm yr⁻¹ initially to 0 cm yr⁻¹ after 25 Myr (Fig. 3B, 3D), resulting
 257 in failed subduction.

258 3.2 Role of Rheology

259 The rheology of the model is a key parameter, as it changes the viscous resistance and therefore
 260 the dissipation partitioning within the domain. The previous models, with all deformation mechanisms
 261 active, are all able to subduct when the slab tip extends to a depth of 150 - 160 km into the mantle (Table
 262 2), with the initial geometry providing the required potential energy to over-come resistive forces at work.
 263 As the critical depth is determined by the resistive forces, we also examined two additional setups that
 264 modified (1) the weak zone decoupling between the subducting and over-riding plate and (2) the removal
 265 of deformation mechanisms operating in the model. Removing deformation mechanisms alters the
 266 strength of the lithosphere and mantle (Fig. 1C), allowing an assessment of the key deformation
 267 mechanisms and key areas of deformation within the system.

268 We find that increasing the coupling between the subducting and over-riding plate, by increasing
 269 the viscosity cap of the weak layer, requires an increase in the initial slab depth, suggesting the fault
 270 interface along convergent margins can inhibit subduction. The required depth of the slab tip for
 271 subduction to develop increases from 125 km (case LS-20-100Y-150) at a viscosity cap of 5×10^{19} Pa s
 272 to 166 km at 1×10^{20} Pa s (case 20-100Y); at 3×10^{20} Pa s (case HS-20-100N-225) subduction is unable

273 to become successful even when the initial slab extends to a depth of ~ 200 km. The results show that the
274 initial potential energy required increases due to an increase in strength of the plate interface, however the
275 maximum duration of the incipient stage remains consistent at < 20 Myr (Fig. 4).

276 The role of deformation mechanisms is also investigated by removing: (i) Peierls mechanism,
277 which operates at higher temperatures than yielding within the slab, reducing the slab's strength and
278 resistance to bending and stretching. (ii) Dislocation creep, which mainly occurs around the slab edges,
279 reducing the mantle strength and thus the drag resistance as the slab subducts. To remove each mechanism,
280 the prefactor for each was modified (Table 1). Both changes in rheology resulted in an increase in the
281 initial slab tip depth required for successful subduction to occur.

282 The removal of dislocation creep results in failed subduction in all cases tested, as the slab is unable
283 to lengthen due to a mantle viscosity of $\sim 10^{21}$ Pa s. This results in diffusion of the slab as no new cold
284 material is advected into the subduction zone. The removal of Peierls creep in NP65-40Y-250 results in
285 successful subduction (Fig. 5) when the slab tip is initially at a depth of 220 km. Both cases NP-65-40Y-
286 250 and NP-65-40Y-275 transition to a successful state early, at around 3 Myr, whilst in NP-65-40N-250,
287 ϕ^{VD} increases above ϕ^{PE} around 17 Myr (Fig. 5). The results show that the duration of the incipient stage
288 remains consistent regardless of rheology due to thermal diffusion of the slab, with a maximum duration
289 of 20 Myr.

290

4. Discussion

291

4.1 The role of viscous dissipation in incipient subduction

292 The energy partitioning during incipient subduction varies greatly over time as the model evolves
293 and the key areas of deformation differs compared to steady-state, successful subduction. The evolution
294 of viscous dissipation over time shows that, in all subducting cases (Fig. 2), the lithosphere (ϕ^{VD}) initially

295 accounts for around 60% of the energy dissipated in the system. This increases in the early stages of
 296 incipient subduction, where the proportion of lithospheric dissipation (ϕ^{VD}) grows to a maximum of
 297 $\sim 80\%$, which is less than the highest estimates of $\sim 95\%$ during steady-state subduction [Conrad and
 298 Hager, 1999; Funiciello et al., 2003; Bellahsen et al., 2005]. This peak in ϕ^{VD} corresponds to the minimum
 299 convergent and slab sinking velocities experienced in the model during its evolution. Although both ϕ_l^{VD}
 300 and ϕ_m^{VD} decrease with decreasing velocity during this stage, ϕ_m^{VD} decreases at a faster rate proportionately
 301 as the lithosphere is undergoing internal deformation, increasing ϕ^{VD} [Capitanio et al., 2007]. This is a
 302 consequence of the need to initially overcome the lithospheric strength, which then allows the instability
 303 to grow subsequently. Once the lithosphere has been sufficiently deformed, ϕ^{VD} begins to decrease and
 304 subduction velocities increase. ϕ_l^{PE} then increases as the instability grows, whilst ϕ^{PE} remains constant
 305 at $\sim 80\%$ as both ϕ_l^{PE} and ϕ_m^{PE} grow proportionally. This critical point, where subduction velocities can
 306 now increase and ϕ^{VD} decreases, defines the point at which subduction has reached a successful state. This
 307 is also observed in other studies, where the peak in the force required to form a subduction zone is due to
 308 overcoming lithospheric resistance [Hall et al., 2003], with an increase in plate age requiring a larger force
 309 for subduction to initiate [Gurnis et al., 2004].

310 In the majority of cases, ϕ^{VD} then decreases to $< 50\%$ as the mantle becomes the main area of
 311 deformation and energy dissipation, in agreement with the lithospheric dissipation values of 35% to 50%
 312 observed by Di Giuseppe et al., [2008]. This highlights the decreasing proportion of energy dissipated
 313 within the lithosphere as subduction evolves [Gerardi et al., 2019]. However, as the majority of cases have
 314 not yet reached a steady-state, due to increasing subduction velocities, values of $\sim 35\%$ [Capitanio et al.,
 315 2007; Gerardi et al., 2019] or as low as $< 20\%$ could be reached [e.g. Schellart, 2009; Stegman et al., 2006;
 316 Capitanio et al., 2009; Stadler et al., 2010]. Values of $\sim 25\%$ observed in 20-20Y as the slab reaches the
 317 660 km transition zone after 60 Myr. However, one case, 100-20Y, shows a sustained high value of ϕ^{VD}

at around 75% for ~ 90 Myr. But, as ϕ^{VD} remains below ϕ^{PE} (Fig. 2E) and the convergence velocity remains high enough, the slabs thermal signature does not diffuse away (Fig. 7A) and subduction continues at a very low rate (Fig. 3A, 3C). Whilst the nature of subduction zones are three-dimensional and the evolution is influenced by toroidal and poloidal flow, the partitioning of energy observed here is similar to 3D models [Stegman et al., 2006], suggesting that the poloidal flow in two-dimension encapsulates the dissipation of potential energy within the system.

4.2 Slab break-off

Slab break-off is observed in 20-20Y (Fig. 6A-B), where the initial rapid acceleration of the subducting slab results in stretching and break-off within 20 Myr (Fig 3A, C). During the evolution of 20-20Y, an initial reduction of viscosity in the slab's core is seen at a depth of around 100 km after 6.4 Myr, where the transition from dislocation to Peierls creep is observed (Fig. 6A-B), whilst thinning of the 1300 K isotherm begins at around 15 Myr. The slab continues to stretch at the necking point, ultimately leading to slab break-off at around 19 Myr (Fig. 6A, 6B).

The slab break-off duration observed here is defined as the duration between the onset of visual necking and complete break-off and occurs over very short periods of time (< 5 Myr) with viscous creep being the dominant deformation mechanism enabling slab break-off (Fig. 6B) in agreement with Duretz et al. [2012]. However, a reduction in viscosity is first observed at 6 Myr, with a time of ~ 13 Myr between the onset of weakening and slab break-off. This suggests that the initial weakening, beginning the slab break-off process, occurs much earlier than visible necking of the slab, suggesting the duration over which slab break-off occurs may be longer than previously thought (Fig. 6A, 6B). However, the duration of slab break-off may be shortened by additional mechanisms, including shear heating [Duretz et al., 2012], grain size reduction [Bercovici et al., 2015] or through strain weakening that records deformation history.

340 The sinking of the detached slab has a two-fold effect: (1) it reduces the viscosity of the mantle
341 through induced strain rates, and (2) it propagates viscous stress, i.e. suction [Baes and Sobolev, 2017;
342 Baes et al., 2018] aiding the continuation of subduction of the remaining slab. This suction force is less
343 than the slab pull, but still aids subduction. The detached slab's thermal signature eventually diffuses away
344 after ~ 27 Myr and subduction of the remaining slab continues. This is observed in an increase in u_y and
345 u_x after ~ 27 Myr until 45 Myr, where velocities begin to decrease as the subducting slab reaches the
346 upper-lower mantle boundary (Fig. 3A, 3C).

347 The weak subducting and over-riding plate in case 20-20Y allows the lithospheric resistance to be
348 easily overcome, resulting in high initial plate velocities (Fig. 3A, 3C). This produces high induced strain
349 rates, reducing the viscosity (Fig. 6), in both the subducting plate, due to bending, and the over-riding
350 plate, due to stretching caused by trench motion, allowing the slab to sink faster when compared with a
351 stronger, thicker over-riding plate, in agreement with previous studies [Yamato et al., 2009; Capitanio et
352 al., 2010; Butterworth et al., 2012]. In case 20-20Y, slab tip velocities (Fig. 3A) are higher than the
353 convergence velocities, i.e. $u_y > u_x$, (Fig. 3C), suggesting the stresses do not fully propagate to the surface,
354 due to tension and stretching of the slab near the surface [Davies and von Blanckenburg, 1995; Gerya et
355 al., 2004]. However, the absolute surface velocities (u_x , Fig. 3A) are low, with the anomalously high tip
356 velocities (u_y , Fig. 3C) arising from the broken slab, suggesting that thermal weakening of the slab [e.g.
357 Duretz et al., 2012; van Hunen and Allen, 2011] at a low Peclet number [Boutelier and Cruden, 2017] aids
358 the slab break-off process. Slab break-off is further aided by a reduction of viscous support of the
359 subducting slab through viscous creep deformation in the mantle (Fig. 6), consistent with previous models
360 [e.g. Gerya et al., 2004; Burkett and Gurnis, 2012]. The plate kinematics and resulting feedback
361 mechanisms, caused by high temperatures around the slab due to the young over-riding plate, results in
362 viscous creep reducing the viscous support of the subducting slab from the surrounding mantle and

increased amounts of deformation at the necking point, leading to slab break-off to occur in case 20-20Y after 19 Myr.

4.3 Constraints on incipient subduction timescales

Our models provide a way to estimate the timescales of incipient subduction, when lithospheric deformation is highest [Gurnis et al., 2004]. The incipient stage is defined as the time until the peak in ϕ^{VD} and subduction velocities are at a minimum. The duration of the incipient stage is < 20 Myr, in agreement with previous estimates on subduction initiation from a range of convergent margins, which are estimated at ~ 10 Myr by Becker et al. [1999], 15 Myr [Faccenna et al., 1999] and up to 20 Myr [Nikolaeva et al., 2010]. Our models, with all deformation mechanisms active, also agree with critical depths that have been suggested in the literature, ranging from 100 – 150 km [Hall et al., 2003; Gurnis et al., 2004] and up to 180 km [McKenzie, 1977].

We find that the age of the subducting plate plays a relevant role in the evolution of incipient subduction margins, whereby increasing the subducting plate age increases the strength of the plate, restricting the bending of the slab. This leads to a longer duration of the incipient stage, where the transition to successful subduction is observed through the ϕ^{VD} peak at ~ 10 Myr and ~ 20 Myr in case 20-100Y (Fig. 2A) and case 100-100Y (Fig. 2C), respectively. This is in agreement with subduction initiation being easier at younger margins as the younger plate requires less force to deform and form the margin [Gurnis et al., 2004; Zhong and Li, 2019].

The role of the over-riding plate in facilitating incipient subduction is also important. A young over-riding plate can accommodate convergence to allow subduction to occur. This is observed in case 100-20Y, with an old subducting plate (100 Myr) and young over-riding plate (20 Myr). In this subducting case, no transition from incipient to successful subduction is observed, with ϕ^{VD} remaining high at $\sim 70\%$ after 80 Myr (Fig. 2E). During the evolution of case 100-20Y, the OP deforms largely as convergence is

ongoing (Fig. 7A), with the high ϕ^{VD} keeping the convergence velocity low. The velocities from the model (Fig. 3C) suggest that stretching of the over-riding plate is facilitating the majority of convergence (Fig. 7A, 7B). The strength of a viscous fluid is very different between bending and extension. During extension, resistance to stretching is proportional to the product of effective viscosity and thickness [Turcotte and Schubert, 2002; Capitanio et al., 2007; Garel et al., 2014] while resistance to bending is proportional to the product of effective viscosity and the cube of the thickness [Turcotte and Schubert, 2002; Ribe, 2001; Capitanio et al., 2007; Garel et al., 2014]. This may be occurring in 100-20Y, where stretching and deformation of the weak over-riding plate facilitates subduction instead of deformation of the subducting plate, limiting the convergence velocity and the advection rate.

The evolution of case 100-20Y, where a weak over-riding plate accommodates the majority of trench motions, suggests that the over-riding plate is an important area of viscous dissipation in the incipient stage, and may be higher than the 0.4% previously suggested during steady-state subduction [Chen et al., 2015]. The slab extends to a depth of ~ 600 km, with ~ 400 km of slab roll-back from the initial trench position, with convergence velocities (u_x) of ~ 0.5 cm yr⁻¹ (Fig. 3C). This suggests subduction can remain in an ongoing, incipient state for long periods. Subduction is actively on-going in case 100-20Y through the lengthening of the slab and penetration into the upper mantle (Fig. 7A-B), where subduction is possible with sustained convergence velocities of ~ 0.5 cm yr⁻¹ (Fig. 3A, 3C). This implies a minimum of ~ 100 km of convergence within 20 Myr for subduction to occur, in agreement with previous studies [Hall et al., 2003; Gurnis et al., 2004]

The failed subduction cases also provide insight into the incipient stage, suggesting that if convergence velocities remain low (< 0.5 cm yr⁻¹), the subducting slab will begin to lose buoyancy due to thermal diffusion, resulting in a further decline in convergence velocity. This leads to less potential energy being released in the system due to its dependence on the slab sinking velocity, u_y (eq. 12). This negative

feedback ultimately leads to failed subduction. In all failed subduction cases (Fig. 2), the lithospheric viscous dissipation exceeds the lithospheric potential energy release rate early in the evolution, between 8 - 22 Myr. Peaks in the lithospheric dissipation are up to 85% in some cases. The high lithospheric dissipation prevents the instability to grow (Fig. 2) and therefore reduces the advection of cold material into the subduction zone. This allows thermal diffusion from the hot mantle to heat up the subducting slab's negative thermal anomaly and therefore subduction fails as the driving force disappears.

4.4 Rheology controls on incipient subduction

The weak zone, at the interface between the subducting and over-riding plate, is a key component required for successful subduction, as an increase in strength increases the plate coupling, resisting the lengthening of the slab length. The weak zone makes a negligible contribution to the lithospheric viscous dissipation throughout the system's evolution if it is weak enough. However, the weak zone can become an important area of dissipation with increasing strength [Conrad and Hager 1999] and may even dissipate similar amounts of energy as the bending zone during steady-state subduction [Stadler et al., 2011]. Our models show that with increasing strength, the influence of the plate interface on subduction and energy dissipation increases. This implies sufficient decoupling is required between the plates, with increasing strength requiring an increase in the initial amount of potential energy. An increase in viscosity from 5×10^{19} Pa s (case LS-20-100Y-150) to 1×10^{20} Pa s (case 20-100Y) increases the required initial slab tip depth for successful subduction by ~ 41 km (125 km to 166 km, respectively). This implies a greater initial potential energy is required for successful subduction, when only the weak zone strength is modified, illustrating the role of the plate interface rheology.

From Fig. 4, different timescales of incipient subduction can be seen based on the initial bending radius, and thus depth. This demonstrates the longer the initial slab, the higher the initial potential energy and the quicker the slab can overcome the lithospheric resistance and accelerate, decreasing the time to

transition from incipient to successful subduction. This is seen in case LS-20-100Y-200 where the lithospheric viscous dissipation peaks after 9 Myr, whilst LS-20-100Y-150 peaks at ~15 Myr.

Effectively removing dislocation creep in the upper mantle highlights the role of mantle drag. This results in diffusion creep becoming the dominant deformation mechanism in the mantle, with a mantle viscosity of $\sim 10^{21}$ Pa s. Diffusion creep around the slab edge is unable to reduce the resistive drag force from the mantle as it cannot reduce the viscosity of the surrounding mantle around the slab as effectively as dislocation creep. Elasticity, although not included here, also influences the evolution of subduction zones. It has been shown to maintain higher deformation rates during bending and retain slab strength during unbending, but has minor effects on the overall morphology [Capitanio and Morra, 2012; Farrington et al., 2014].

The removal of Peierls mechanism highlights the role of the strength of the slab. In case NP-65-40Y-250, yielding of the slab is able to facilitate some deformation and reduce the viscosity of the slab. However, yielding is not as efficient as Peierls mechanism at reducing the viscosity in those regions, increasing the amount of potential energy required to reach successful subduction. However, the duration of the incipient stage remains consistent at < 20 Myr (Fig. 5).

The variations in rheology investigated here suggests that both lithospheric and mantle deformation are key components for successful subduction to occur, similar to observations by Jadamec [2015]. In all cases that evolve successfully, the transition from incipient to successful subduction varies in time, but are all within 20 Myr. The model's initial bending radius and rheology can alter the evolution, with an increase in the length of slab increases the slabs potential energy, decreasing the duration of the incipient stage. However, if the lithospheric strength cannot be over-come and the instability cannot grow within 20 Myr, the slab thermally diffuses away and subduction fails, regardless of the local conditions.

454 4.5 Topography constraints on incipient subduction

455 The free surface in the model allows an assessment of the topography at convergent zones during
456 incipient subduction. As the topography is the surface expression of large-scale geodynamic processes,
457 this allows a first-order comparison between incipient subduction zones and the model evolution [Crameri
458 et al., 2017].

459 Each case has a unique topographic evolution, providing useful diagnostics. However, there are
460 common features across all successful cases. During the incipient stage, the subduction zones are
461 characterised by: (1) an area of subsidence in the bending region of the subducting slab up to 200 km wide
462 and (2) subsidence in the over-riding plate due to slab roll-back, or a back-arc depression (Fig. 8A), as
463 illustrated by Crameri et al. [2017]. Successful subduction is characterised by additional features (Fig.
464 8C), including: (1) the formation of a fore-bulge, with uplift in the “outer-rise” up to 200 m above the
465 original free surface, (2) the presence of an extensional basin where the main area of slab bending is
466 occurring, (3) a trench up to 6 km deep, similar to topography in Toth and Gurnis [1998]. Most cases also
467 show an area of uplift in the over-riding plate prior to the back-arc depression, similar in appearance to an
468 island arc [Crameri et al., 2017]. In failed cases, these features do not develop and only a proto-trench,
469 where the initial perturbation was present, is visible, at < 4 km deep, that narrows as subduction ceases in
470 the region (Fig. 8B, 8D).

471 These results allow a direct comparison with topography in nature. Although lack of extensive
472 observations prevents a thorough comparison, our work is compatible with reported cases. The Vitiaz
473 trench, as discussed by Gurnis et al. [2004] is a fossil subduction zone, where subduction ceased ~ 10 Ma.
474 The topography at the Vitiaz trench (VT) is very similar to that seen in the failed subduction cases, where
475 the trench is still visible as a depression in the surface (Fig. 9B), however it is not as deep (~ 1 km), as
476 surrounding active subduction zones, such as the New Hebrides trench (NHT) (Fig. 9B). The topography

of the NHT shows the same features observed in successful subduction cases, with uplift in the fore-bulge and island-arc region either side of the trench, as well as some subsidence in the active bending area.

In comparison, the Macquarie Ridge Complex, off of the southwest coast of New Zealand, was defined by Gurnis et al. [2004] as a ‘forced’ subduction zone. Topography shows (1) a well-defined trench ~30 km wide, the Puysegur Trench; (2) uplift in the back-arc region, the Puysegur ridge and (3) development of a trough between 10 km to 40 km wide on the over-riding plate between 47°S and 48°S (Fig. 9A), indicative of slab roll-back occurring in that region. There is also some subsidence prior to the trench, however it is not as well defined as subsidence in the NHT, and no fore-bulge is observed in the topography. Although the area is complicated by oblique convergence, the topography in the region suggests that the Macquarie Ridge Complex could be considered to have reached a self-sustaining, successful state in some areas.

5. Conclusions

We have studied the incipient stage of subduction within a thermomechanical system through the evolution of a short slab into a fully self-sustained subduction in the upper mantle. During incipient subduction, the available forces are accommodated mostly by internal deformation of the down-going and over-riding plate. This inhibits subduction development, which slows down slab velocities and causes diffusion of the slab’s thermal buoyancy, further decreasing driving forces. This negative feedback may ultimately lead to subduction failure. We assess the evolution of this stage through energy dissipation and find that the critical time for subduction to develop from incipient to successful subduction is < 20 Myr, with average instability growth rates of $\geq 0.5 \text{ cm yr}^{-1}$, resulting in a minimum of 100 km of convergence required. In all failed cases, we observe the lithospheric viscous dissipation (Φ^{VD}) exceeding the

498 lithospheres potential energy release rate (ϕ^{PE}) early in the evolution, < 20 Myr, ceasing subduction due
499 to a lack of driving forces caused by thermal diffusion of the slab.

500 The incipient phase is characterised by low convergence velocities and a high proportion of
501 lithospheric dissipation of available potential energy. The age of the over-riding plate influences the
502 duration of the incipient stage. Increasing the age of the over-riding plate decreases plate velocities, as
503 they are rigid, restricting trench motion. Young over-riding plates increase trench motion; however, they
504 can also hinder subduction velocities in extreme cases. A young over-riding plate can accommodate the
505 majority of deformation through stretching, whilst the subducting plate remains relatively less deformed
506 and unable to subduct freely, resulting in low convergence rates and a slow evolution of the margin.

507 The rheology of both the mantle and lithosphere also influence the dissipation partitioning and,
508 therefore, the incipient subduction stage. In the lithosphere, yielding and Peierls creep critically decrease
509 the resistance to bending at the hinge, key area of deformation required to reach a successful state. In the
510 mantle, dislocation creep, driven by slab-induced flow, allows sufficient deformation around the slab to
511 reduce drag resistance. Without these mechanisms operating, a significant increase in the amount of
512 potential energy is required to reach a successful state. Similarly, an increase in the strength of the
513 subduction zone interface, increasing the inter-plate coupling, requires a larger initial potential energy.
514 These results highlight the key areas of deformation which are required to reach a successful state.

515 The models show distinctive topographic evolution, allowing a clear discrimination between
516 successful and failed subduction. Although the record of failed subduction is scarce, a comparison
517 between reported observed cases and models provides a positive match, showing insights in a process that
518 might be relatively poorly represented, yet important for our understanding of subduction zone dynamics.

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701 Table 1: Model parameters.

Quantity	Symbol	Units	Value
Gravity	g	m s^{-2}	9.8
Coefficient of thermal expansivity	α	K^{-1}	3×10^{-5}
Thermal diffusivity	κ	$\text{m}^2 \text{s}^{-1}$	10^{-6}
Reference density	ρ_s	kg m^{-3}	3300
Cold, surface temperature	T_s	K	273
Hot, mantle temperature	T_m	K	1573
Gas constant	R	$\text{J K}^{-1} \text{mol}^{-1}$	8.3145
Maximum viscosity	η_{max}	Pa s	10^{25}
Minimum viscosity	η_{min}	Pa s	10^{18}
Weak zone viscosity cap ^f	η	Pa s	10^{20}
<i>Diffusion Creep</i>			
Activation energy	E	kJ mol^{-1}	300 (UM)
			200 (LM)
Activation volume	V	$\text{cm}^3 \text{mol}^{-1}$	4 (UM)
			1.5 (LM)
Prefactor ^b	A	$\text{Pa}^{-1} \text{s}^{-1}$	3.0×10^{-11} (UM)
			6.0×10^{-17} (LM, $\therefore \Delta\eta = 30$)
	n		1
<i>Dislocation Creep (UM)^c</i>			
Activation energy	E	kJ mol^{-1}	540
Activation volume	V	$\text{cm}^3 \text{mol}^{-1}$	12
Prefactor	A	$\text{Pa}^{-n} \text{s}^{-1}$	5.0×10^{-16}

	n		3.5
<u>Peierls Creep (UM)^c</u>			
Activation energy	E	kJ mol ⁻¹	540
Activation volume	V	cm ³ mol ⁻¹	10
Prefactor	A	Pa ⁻ⁿ s ⁻¹	10 ⁻¹⁵⁰
	n		20
<u>Yield Strength Law</u>			
Surface yield strength	τ_s	MPa	2
Friction coefficient	f_c		0.2 ^d
	$f_{c, \text{weak}}$		0.02 (weak layer) ^f
Maximum yield strength	$\tau_{y, \text{max}}$	MPa	10,000 ^e

- A. UM and LM are acronyms for “upper mantle” and “lower mantle,” respectively.
- B. We use a generic prefactor with no explicit dependency on water content and grain size, assumed constant in the model.
- C. To reduce non-Newtonian creep in the lower mantle, and upper mantle in additional cases, we set minute prefactors of 10^{-42} and 10^{-300} for dislocation and Peierls creeps, respectively.
- D. A friction coefficient of 0.2 is intermediate between lower values of previous subduction models [Di Giuseppe *et al.*, 2008; Crameri *et al.*, 2012] and the actual friction coefficient of the Byerlee law [Byerlee, 1978].
- E. A very high value is taken for cases with Peierls mechanism, for which the yield mechanism dominates only at shallow depth.
- F. Modified in additional cases.

Table 2: Table of cases, where plate age and geometry were changed, cases with a Y denote subducting cases and N denote failed subduction cases. The transition time represents either: (a) the transition incipient to successful subduction in subducting cases; or (b) the point at which the proportion of lithospheric viscous dissipation exceeds the lithospheric potential energy release rate in failed subduction cases.

Case	Subduction	SP Age	OP Age	Bending	Bending	Initial	Transiti	Modific
Name				Radius	Angle	Depth	on Time	ations
						below		
		(Myr)	(Myr)	(km)	(°)	surface	(Myr)	
						(km)		
20-20Y	Y	20	20	200	77	166	8 ^a	-

20-20N	N	20	20	200	67.5	141	10.5 ^b	-
20-100Y	Y	20	100	200	77	166	10 ^a	-
20-100N	N	20	100	200	67.5	141	10 ^b	-
100-100Y	Y	100	100	200	67.5	162	20 ^a	-
100-100N	N	100	100	250	45	145	6 ^b	-
100-20Y	Y	100	20	200	67.5	162	-	-
100-20N	N	100	20	250	45	145	20 ^b	-
65-40Y	Y	65	40	190	77	185	12 ^a	-
65-40N	N	65	40	200	67.5	154	20 ^b	-
40-65Y	Y	40	65	170	77	150	12 ^a	
40-65N	N	40	65	140	77	124	12 ^b	
LS-20- 100Y-200	Y	20	100	200	77	166	7 ^a	WZ = 5x10 ¹⁹ Pa s
LS-20- 100Y-150	Y	20	100	150	77	125	15 ^a	WZ = 5x10 ¹⁹ Pa s
LS-20- 100N-125	N	20	100	125	77	104	5 ^b	WZ = 5x10 ¹⁹ Pa s
HS-20- 100N-225	N	20	100	225	77	202	-	WZ = 3 x 10 ²⁰ Pa s

NP-65-40Y-275	Y	65	40	275	77	246	2.5 ^a	No Peierls Creep
NP-65-40Y-250	Y	65	40	250	77	220	2.5 ^a	No Peierls Creep
NP-65-40Y-225	N	65	40	225	77	202	18 ^b	No Peierls Creep

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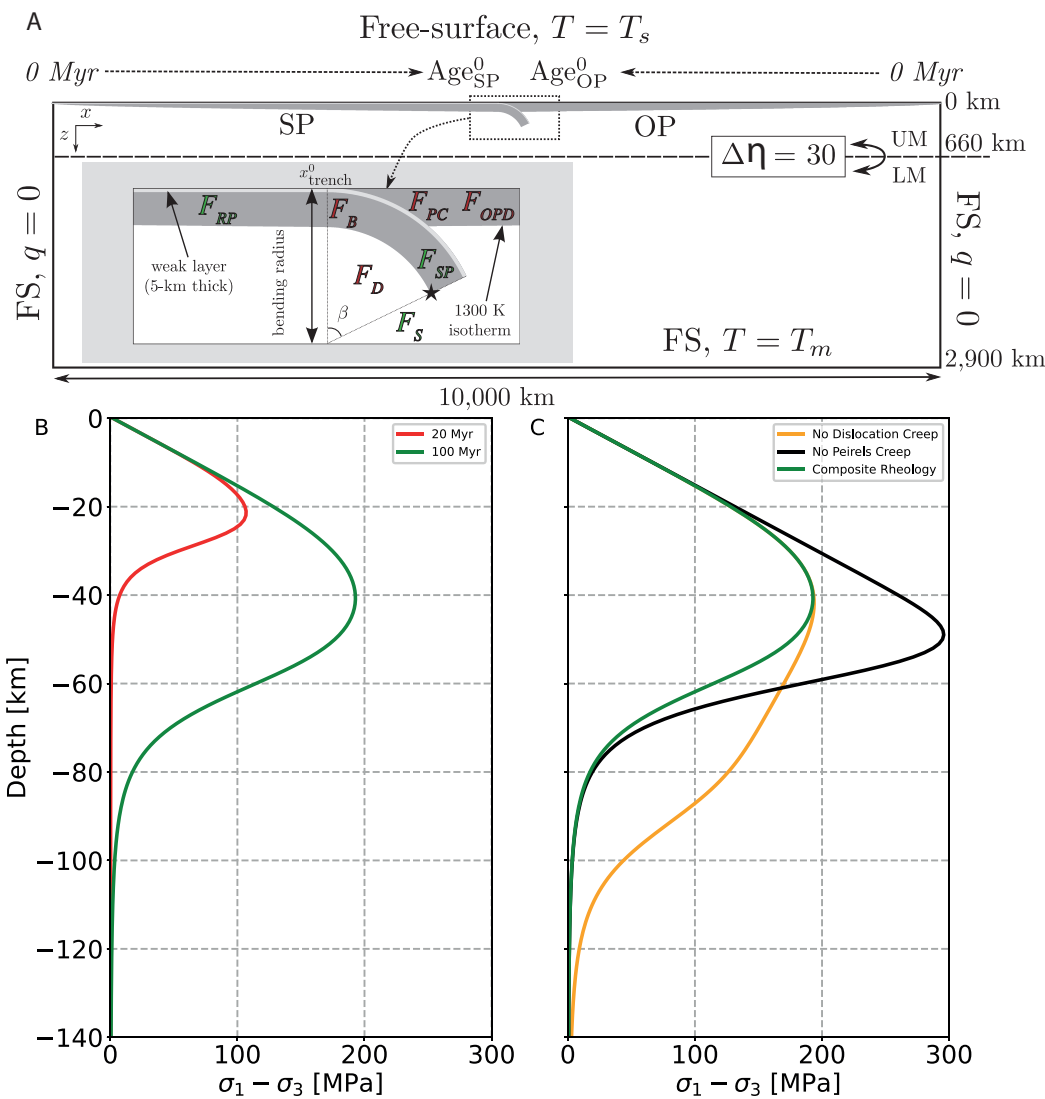


Fig. 1: A) Initial model setup, modified from Garel et al. [2014]. No rheological distinction is made between slab and upper mantle. Heat flux, q , is 0 at the side boundaries. Free-slip (FS) boundary conditions are imposed on the bottom and sides, with a free surface along the top. Age^0_{SP} and Age^0_{OP} are the initial ages of the subducting (SP) and overriding plates (OP) at the trench, respectively. $\Delta\eta = 30$ is the jump between diffusion creep upper (UM) and lower mantle (LM) viscosity at 660 km. The initial hook geometry of the subducting plate is prescribed using a variation in bending radius (including the weak layer, shown in light grey) and β . The star indicates the location at which the initial slab tip depths have been measured. The driving forces (green) – Ridge push (F_{RP}), slab pull (F_{SP}) and suction due to slab break-off (F_S) as well as resisting forces (red) – bending (F_B), mantle drag (F_D) plate coupling (F_{PC}) and over-riding plate deformation (F_{OPD}) are included. Modified from Garel et al. [2014]. B) Initial strength profiles using a composite rheology and a background strain rate of 10^{-15} s^{-1} of a 20 Myr and 100 Myr slab at the trench. C) Initial strength profiles of a 100 Myr slab at the trench with a composite rheology, or with dislocation or Peirels creep removed and a background strain rate of 10^{-15} s^{-1} .

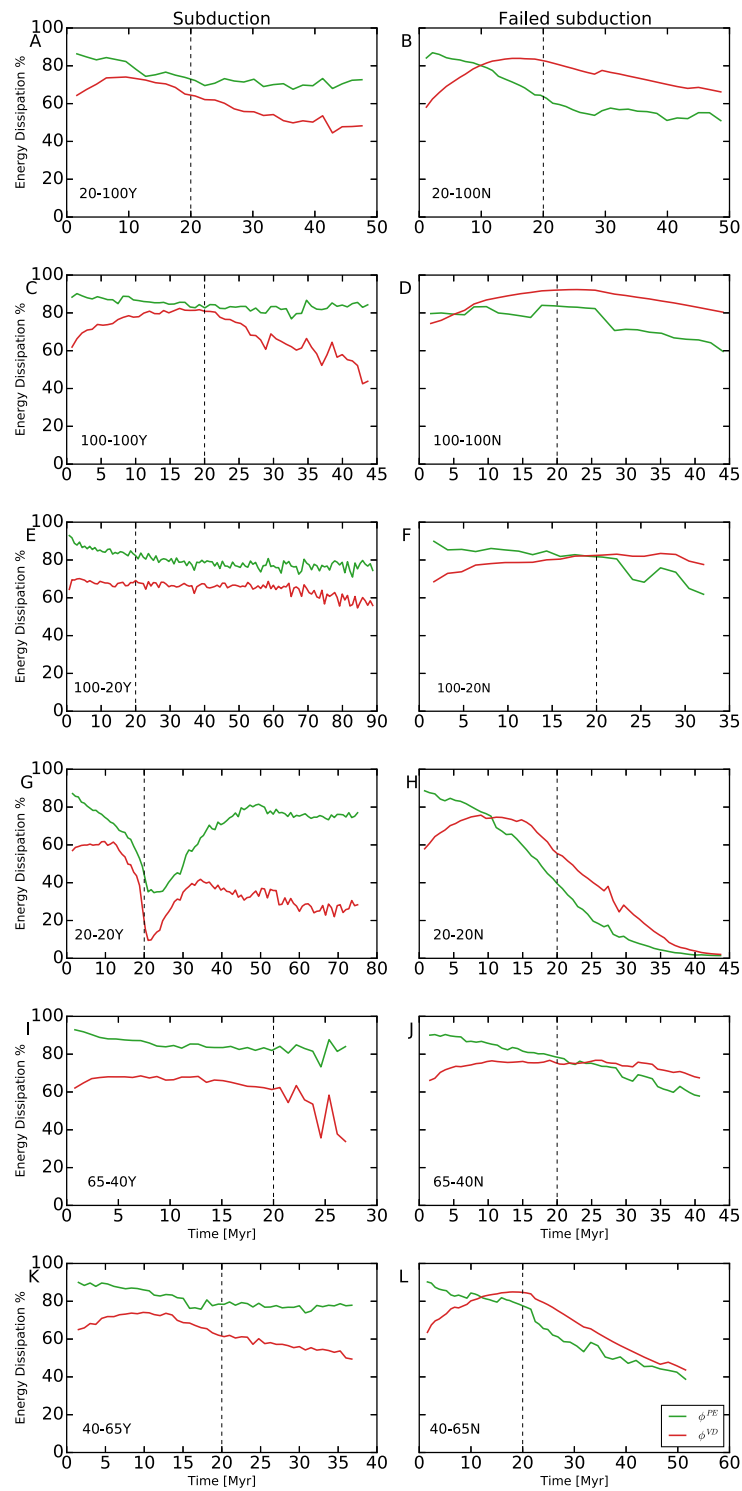
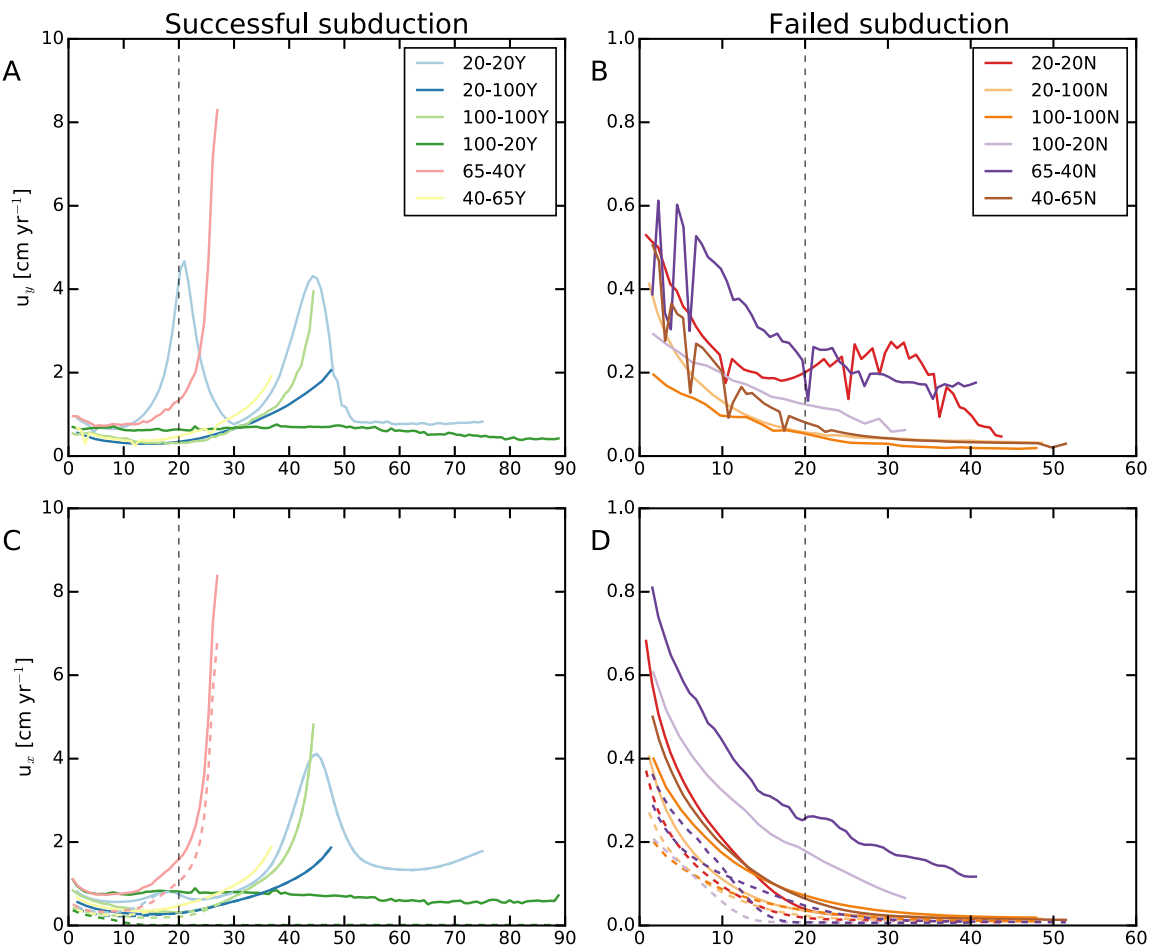


Fig. 2: Evolution of dissipation of lithospheric potential energy release rate and viscous dissipation for each model, case numbers are included in the panel. The red line represents the percentage of lithospheric viscous dissipation in the model (ϕ^{VD}) whilst the green line represents

733 the percentage of lithospheric potential energy release rate (ϕ^{PE}). The dotted line represents 20 Myr in the evolution, the critical duration
734 of the incipient stage. All failed cases have ϕ^{PE} decrease below ϕ^{VD} within 20 Myr, which is not observed in successful cases.



735
736 Fig. 3: A) Vertical velocity (u_y) of the slab tip of successful subducting cases. B) Vertical velocity (u_y) of the slab tip of failed subduction cases.
737 C) Convergence velocity (u_x - solid) and subducting plate horizontal velocity (u_{sp} - dashed) of successful cases. D) Convergence velocity (u_x -
738 solid) and subducting plate horizontal velocity (u_{sp} dashed) of failed cases. All Subducting cases have a minimum convergent velocity of ~
739 0.5 cm yr^{-1} , whilst all failed cases tend toward 0 cm yr^{-1} .

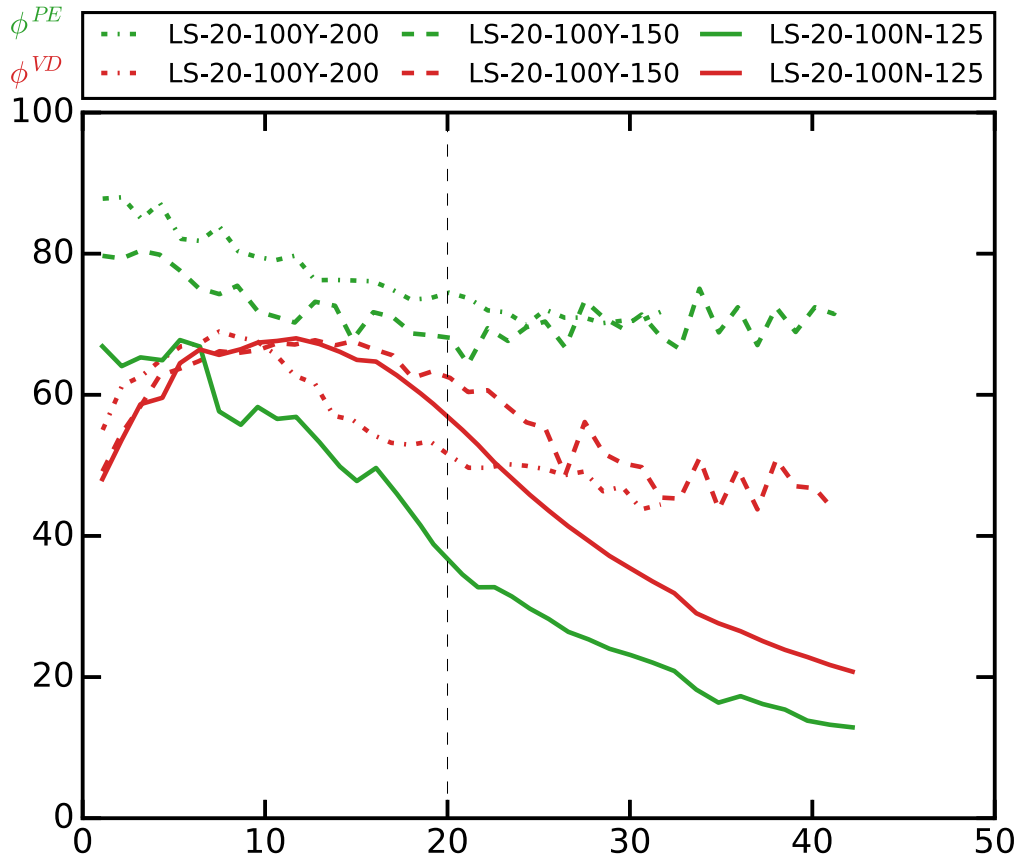


Fig. 4: Energy dissipation when a viscosity cap of $5 \times 10^{19} \text{ Pa s}$ is applied to the weak zone, with varying bending radius (BR), and thus initial slab tip depth below surface (IDBS), for cases LS-20-100Y-200, LS-20-100Y-150 and LS-20-100N-125 (BR – 200, 150, 125 km, IDBS – 166, 125, 104 km, respectively). The results suggest that subduction reaches a successful for a bending radius of $> 150 \text{ km}$ (LS-20-100Y-150) or an initial slab tip depth of 125 km. The results also show that with an increasing slab length, the lithospheric strength is easier to overcome and the duration of the incipient stage is shorter due to the slabs higher potential energy.

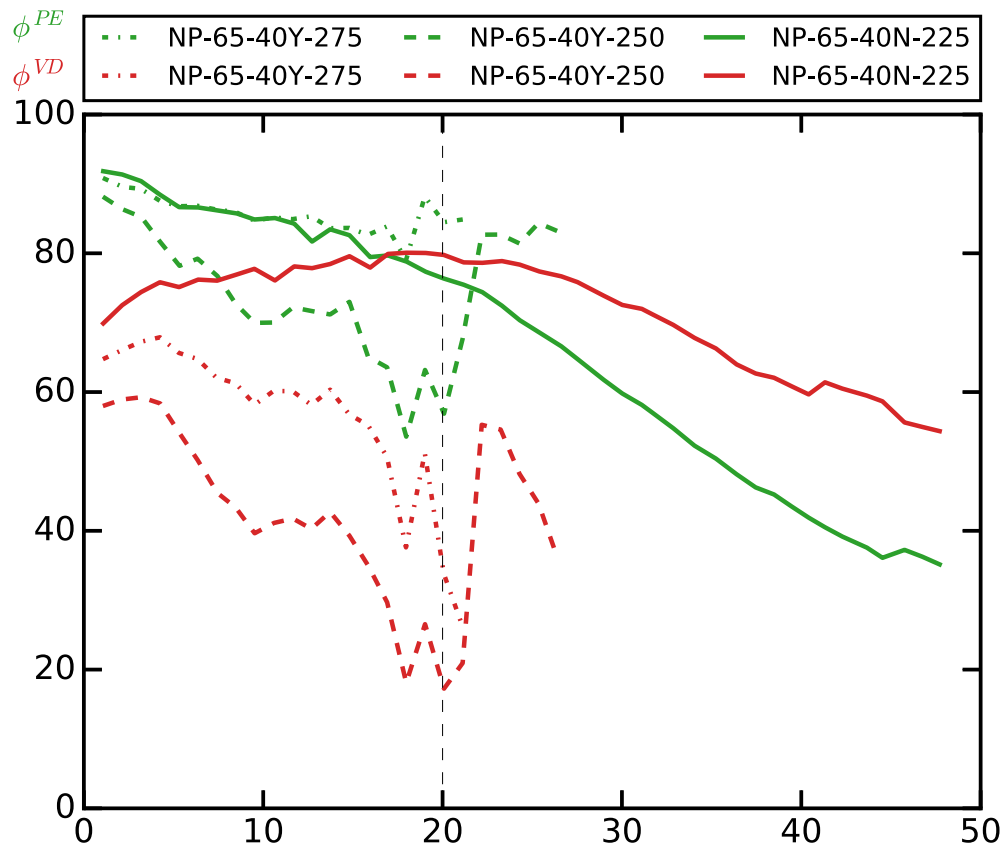


Fig. 5: Energy dissipation of the case without the Peierls creep deformation mechanism, with varying bending radius (BR), and thus initial slab tip depth (IDBS), for cases NP-65-40Y-275, NP-65-40Y-250 and NP-65-40N-225 (BR – 275, 250, 225 km, IDBS – 246, 220, 202 km, respectively). The results show an initial BR of 250 km (depth of >220km, case 65-40Y-B) is required for subduction to reach a successful state. In these cases, the lithospheric viscous dissipation peaks early and transitions to mantle dominated as the slab remains strong and the mantle is preferentially deformed due to the removal of the Peierls creep deformation mechanism.

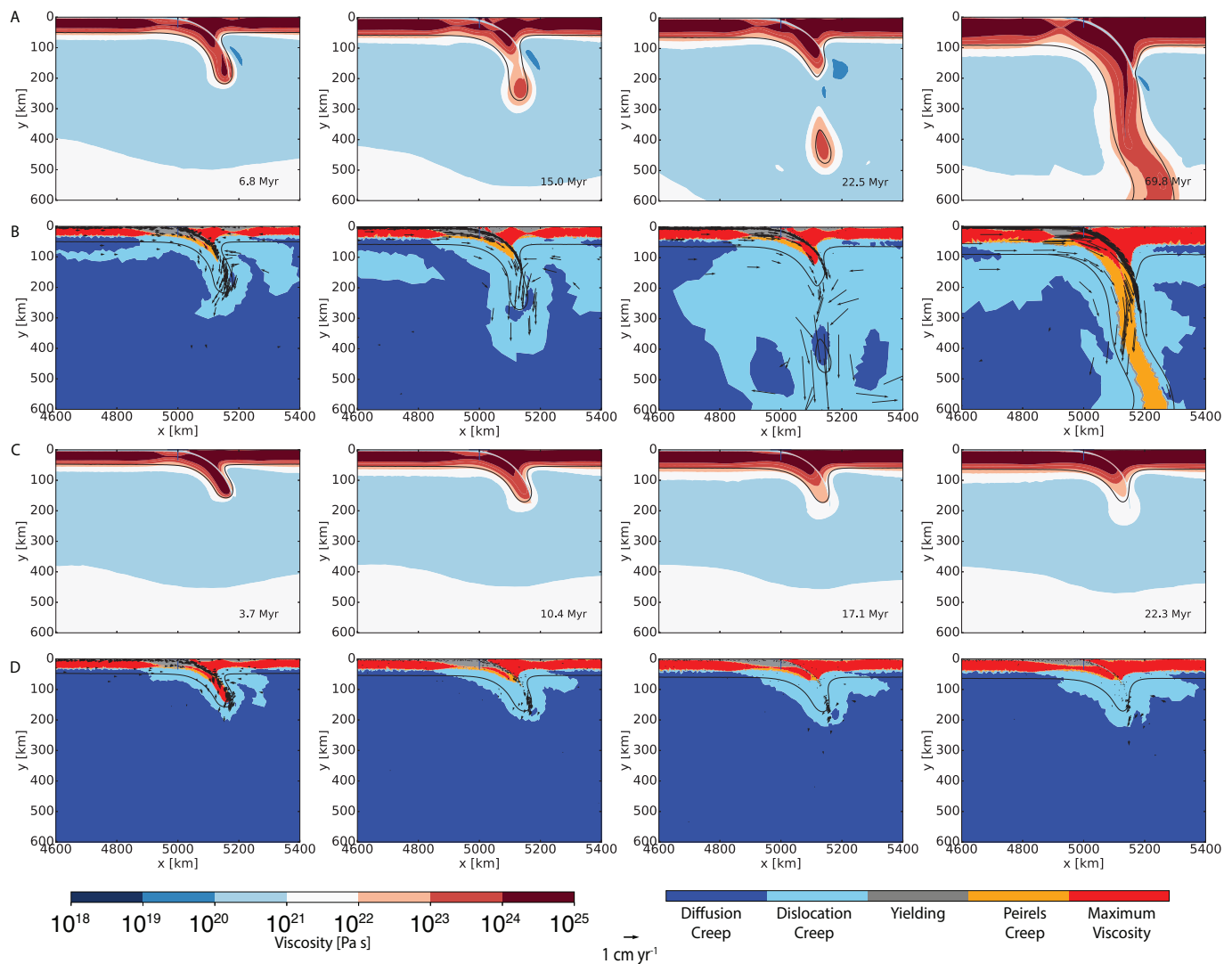


Fig. 6: Evolution of case 20-20Y showing a) Viscosity and b) main deformation mechanism with velocity vectors and 20-20N showing c) Viscosity and d) main deformation mechanism with velocity vectors. The effect of slab break-off is seen clearly in the mantle where lower levels of viscosity and suction, highlighted by the velocity vectors, are caused by the broken portion of the slab as it sinks freely in the upper mantle. Case 20-20N shows how the slab thermally diffuses away over 20 Myr.

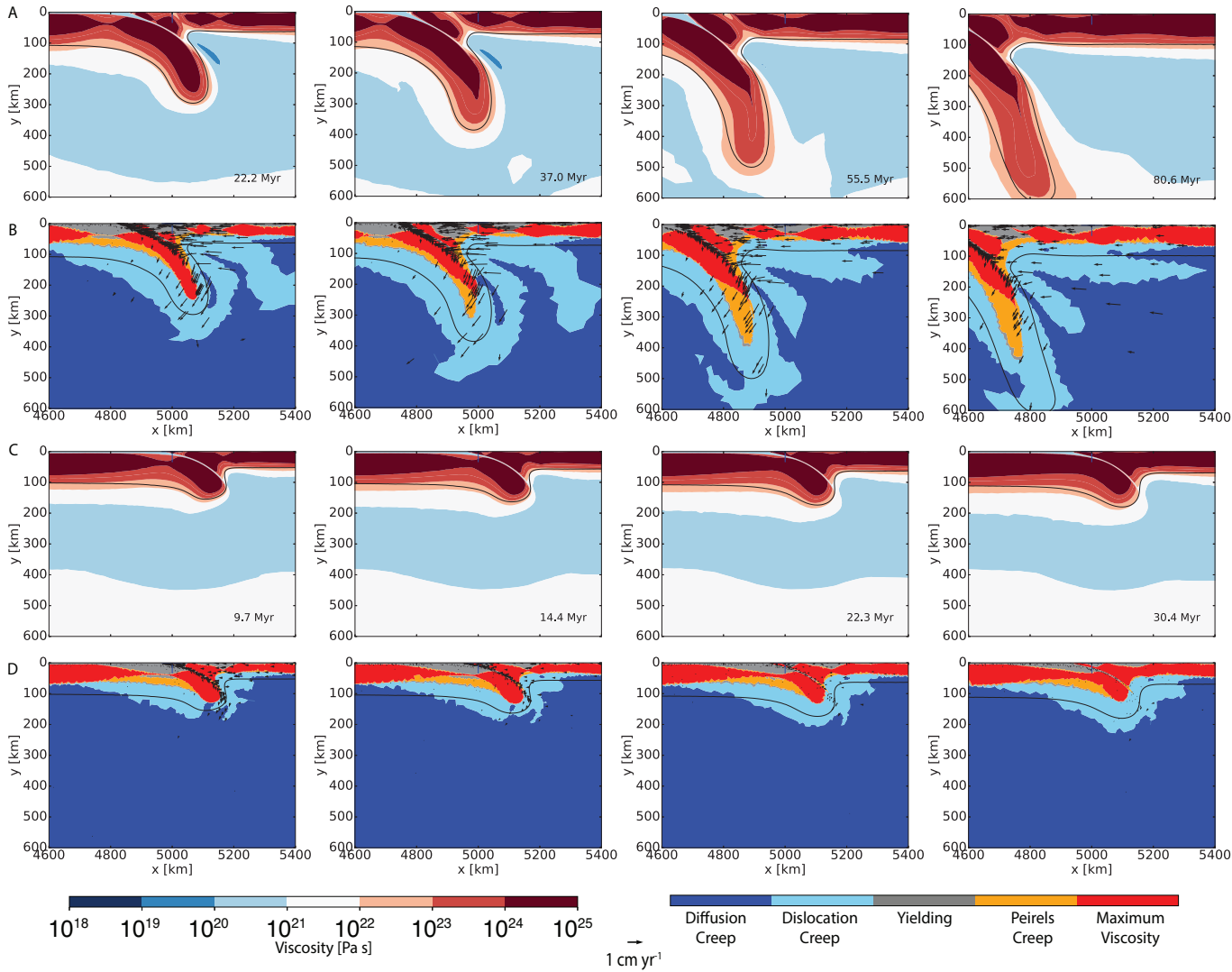


Fig. 7: Evolution of case 100-20Y, showing a) Viscosity and b) main deformation mechanism and velocity vectors and case 100-20N, showing c) Viscosity and d) main deformation mechanism and velocity vectors. The evolution of case 100-20Y demonstrates the deformation and reduced strength in the over-riding plate caused by slab roll-back, and the subducting plate remains rigid. 100-20N shows how the slab thermally diffuses away over 20 Myr.

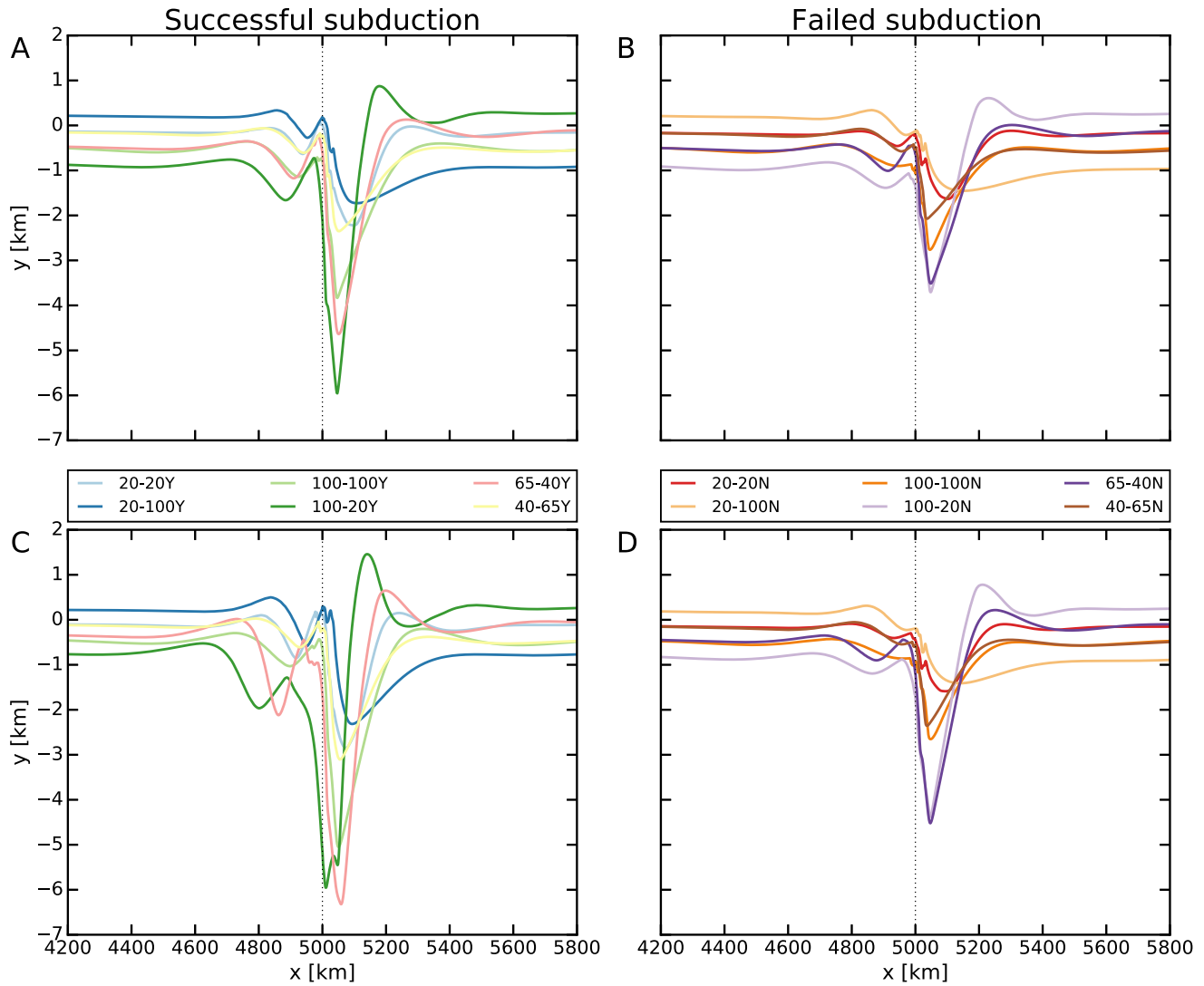


Fig. 8: Free surface heights for each case, representing topography. A) Free surface after 5 Myr for successful cases, with the development of topographic features. B) Failed cases after 5 Myr, development of some topography, but does not develop further. C) Subducting cases, with larger extremes in free surface height associated with older subducting slabs, however the wavelength of each feature remains fairly consistent across a range of plate ages. D) Failed subduction cases, with the trench still clearly visible in each case.

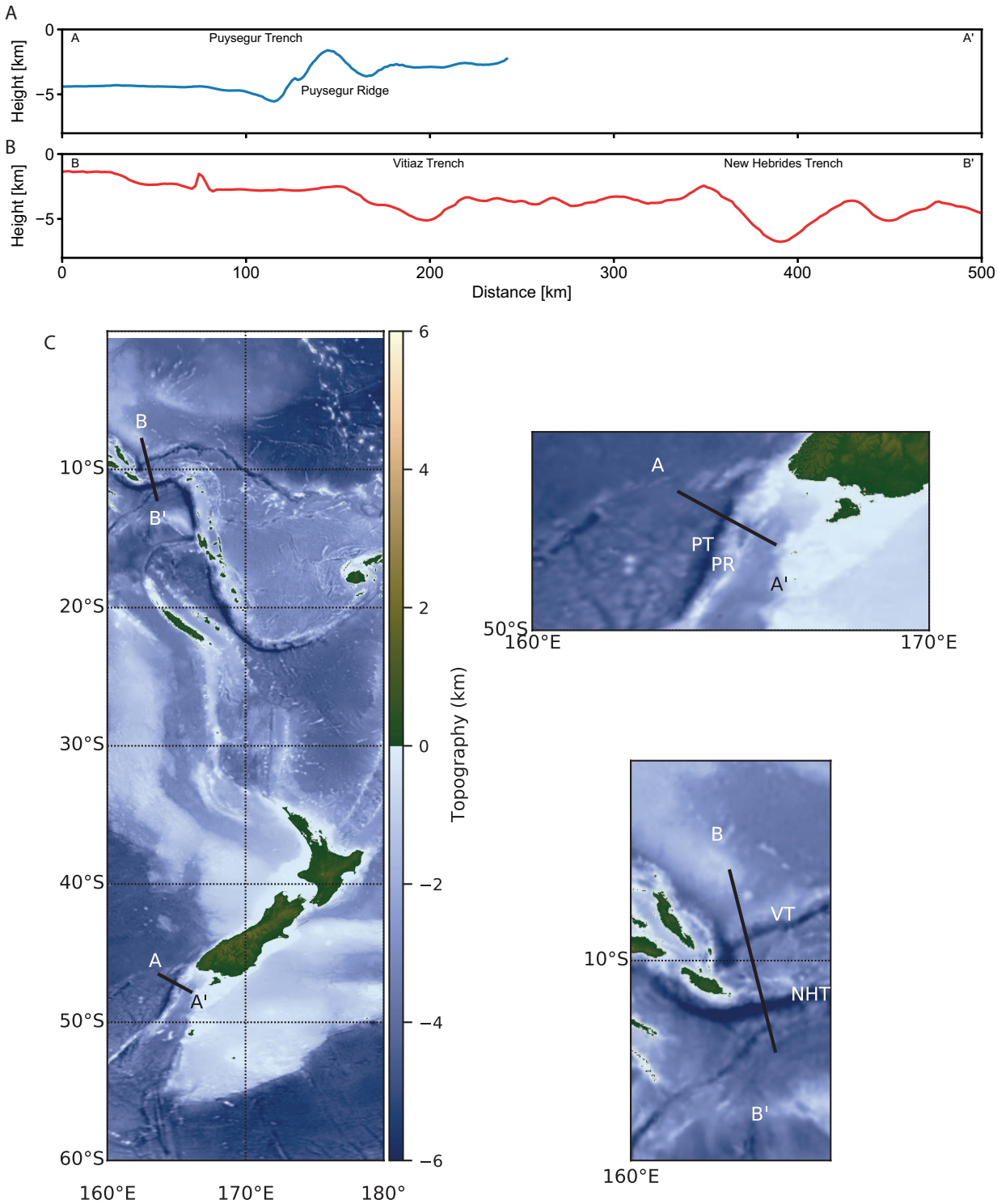


Fig. 9: A) Topography across the Psysegur subduction zone. B) Topography across the Vitiaz (VT) and New Hebrides (NH) trenches. C) Topographic data from etopo1 [Amante and Eakins, 2009].